Quasi-Newton algorithm using Fresnel wavepaths and frequency increase for P-wave tomography inversion: application to a landslide in the South French Alps

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Problematic

During last decases, geophysical methods have become of a great interest in geomorphological studies. Because they are well adapted to retrieve geological structures as variations in the spatial and temporal dimensions of rocks properties, they were widely developed for improving landslides understanding. Landslide studies generally involve the use of several geophysical methods, but among them, seismic surveys are well adapted to identify the slope's main structures. The wave propagation being mainly controlled by elastic properties of the medium, this method makes the interpretation easier since results are often well correlated with geotechnical observations. More generally, it provides information on the mechanical state of the soils with an acceptable spatial resolution.

This structure is of first importance when studying clayey landslides as the Super-Sauze one. It occurred in the 1960s with the falls of large blocks and has developed continually covering an intact paleotopography. This succession of crests and gullies has been studied by geotechnical measurement and geophysics. It plays a large role in the behavior of the flow by delimiting preferential water and material pathways and compartments with different kinematics, mechanical and hydro dynamical characteristics. For the first time, a 3D geological model has been created from the fusion of multi-source data by Travelletti and Malet (2011), but it appears that geophysical methods can't capture the sharp geometry of the paleotopography.
To improve such models and the numerical modeling resulting we propose a Quasi-Newton algorithm based on the Fresnel-wavepath and the frequency increase to the invert P-wave velocity field

Theoretical approach

We choose to use the concept of sensitivity kernels to improve the resolution of the tomography. While the ray theory is well adapted in media with structures scale larger than the first Fresnel zone, the use of sensitivity kernel permits to overcome this constraint and then, to increase the resolution (Spetzler and Snierder, 2004).

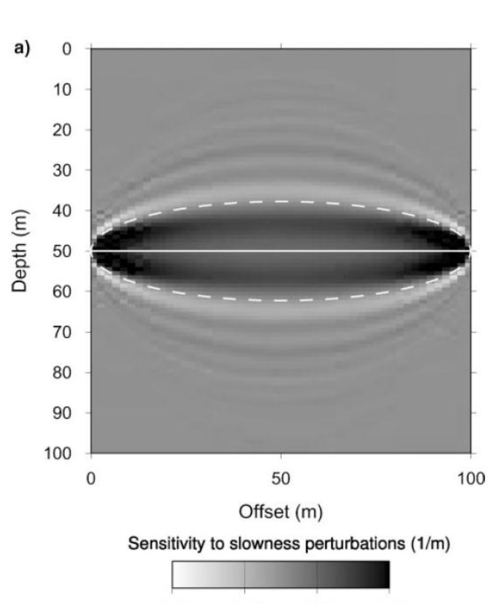

Figure 1 : 2D traveltime sensitivity kernel for transmitted wave (Spetzler and Snieder, 2004)

Sensitivity kernels are based on single scattering approximation, also called Born. It considers that a part of the wave can be delayed by a scattering object surrounding the ray path (Fig. 1). In the particular case of highly heterogeneous media, this assumption is not applicable anymore because of the importance of multiple scattering. That's why, we consider that the high heterogeneity of the soil around the ray path affects through complex multiple scattering the first arrival of the signal. In other word, we give up the single scattering approximation, considering the complex multiple scattering empirically inside the first Fresnel zone.

We define the Fresnel weights proposed by Watanabe et al. (1999) and used by Grandjean and Sage (2004) to materialize the sensitivity kernel. This weights are also used to retropropagate residuals in the update slowness vector according to a Quasi-Newton algorithm. They classically decrease linearly from 1 (when the cell is positioned on the ray path) to zero (when it is out of the Fresnel volume):

$$\omega = \begin{cases} 1 - 2f\Delta t, & 0 \le \Delta t < \dfrac{1}{2f} \\ 0, & \dfrac{1}{2f} \le \Delta t \end{cases}$$

Using such statements, the traveltime perturbation can be expressed as the integral over a Fresnel volume multiplied by the slowness perturbation field over all points r in the volume (Liu et al, 2009). This linear relation between traveltime and slowness perturbation is the result of the first Born approximation only valid for small perturbation (Yomogida, 1992):

$$\delta t = \int W(r)\delta s(r)dr$$

Here, W is the Fresnel weight operator, normalized by the area of the Fresnel ellipse perpendicular to the raypath α, for each shot and for each receiver such as :

$$\int_{-\infty}^{+\infty} W(s,r,f)ds = 1$$

The Fresnel volume thus defined, also called Fréchet kernel correspond to the Fréchet derivative or Jacobian matrix of the forward problem (Yogomida, 1992), similar to the length of ray segment matrix L defined in the asymptotic high frequency assumption. So that the inversion problem considered can be solved with a Quasi-Newton algorithm by solving this linear tomographic system with the LSQR algorithm :

$$\left[\left(W^k\right)^T C_D^{-1}\left(W^k\right)\right]\delta m = \left[\left(W^k\right)^T \delta t\right]$$

Because the size of the Fresnel volume thus defined is dependent of the source frequency considered, a new inversion strategy based on increasing frequency can be evaluated. We propose to compute the Fresnel weights for a monochromatic wave, increasing its frequency at each step of the inversion. For low values, the Fresnel zone will be wider and the slowness model will be reconstructed over a large area around the theoretical wavepath. Conversely, for high frequencies, the slowness model will highlight sharper zones. This strategy appears efficient to improve the resolution during the inversion and also the convergence of the algorithm.

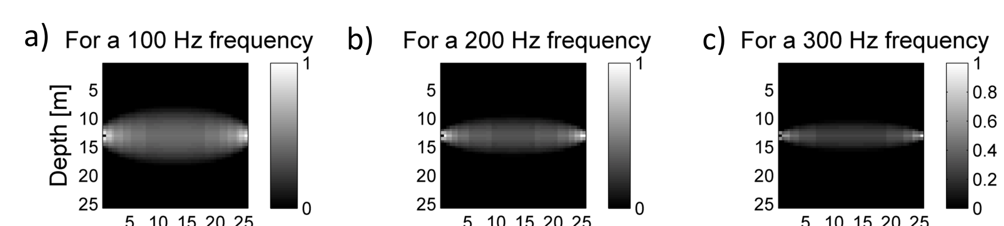

Figure 2 : Fresnel weights defined by Watanabe (1999) for different frequencies (Top) and sensitivity kernels used in the Quasi-Newton algorithm computed for different frequencies (Bottom)

Application to the Super-Sauze landslide

The previously described algorithm was applied on a real dataset acquired on the Super-Sauze landslide. The seismic profile was parallel to the flow axe (longitudinal profile). It was situated in the upper part of the landslide from the second escarpment (transect A, see Malet, 2003), to the middle of the earthflow (transect D, see Malet, 2003). The seismic antenna was 238 m long and has 120 geophones regularly spaced every 2 m. The 60 shots have been achieved with a hammer every 4 m. We used a roll-along system to translate the flute and recorded only 48 traces per shot. Only the vertical component of the seismic signal was measured. The initial length of recording is 1.5 s with a sample rate of 0.25 ms with a Geometrics Stratavisor seismic camera with 48 geophones of central frequency 10 Hz. After picking, differences between reciprocal traveltimes have been computed and picks with a difference greater than 10 ms have not been used for the inversion. The error in reciprocity of traveltimes can be approximated by a Gaussian belt with a standard deviation of 2.2 ms (Fig.3).

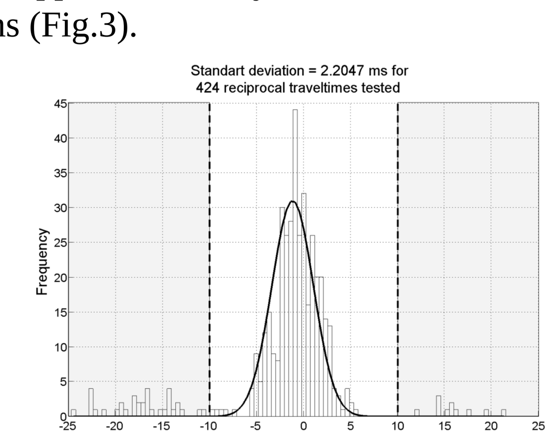

Figure 3 : Errors in traveltime reciprocities

The signal is dominated by the surface wave signal but the first arrivals are clearly visible. The inversion has been performed with the QN algorithm on a 358x153 grid. Each square cell of the grid measures 0.67 m. The hammer source gave frequencies comprised between 30 Hz and 120 Hz with a dominant frequency of 40 Hz.
Those values were considered as constant for each shot. For the first iteration, the frequency was set to 30 Hz and increases at each step to 45 Hz, 60 Hz, 75 Hz, 90 Hz, 105 Hz, until 120 Hz, so that after the eighth step, the frequency was kept constant.

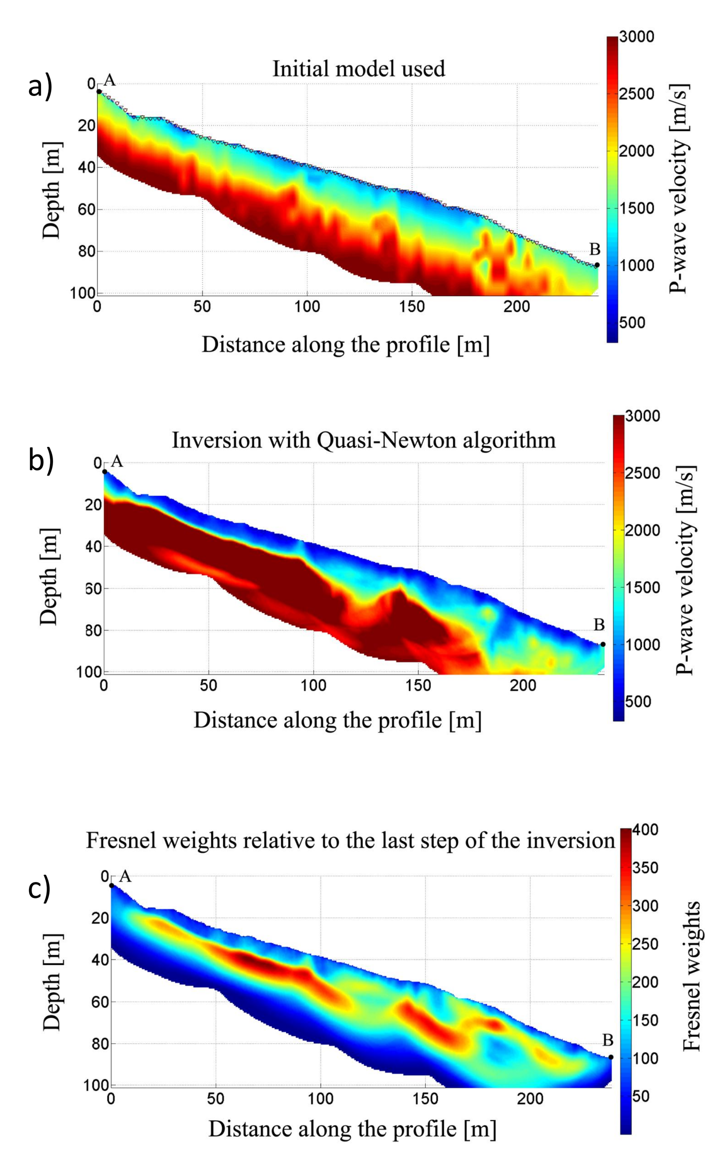

Figure 4 : a) Initial model constructed automatically from the dataset. b) Inversed model with Q-N algorithm c) Fresnel weights relative to the last step of the inversion show the sensibility of the model

Seismic wave attenuation

The Super-Sauze landslide offers a promising context to study the effect of fissuring on seismic wave attenuation. The amplitude of the first arrival was thus exploited to recover the lateral variation of alteration state of the flow layer. The seismic wave attenuation is a good physical property because it is directly linked to porosity and to the presence of fractures in the soil. We choose to use the simplest inversion method proposed by Watanabe and Sassa (1996) considering that in a homogeneous attenuation media the amplitude of the spherical wave verifies:

$$A(r) = \frac{A_0}{r}e^{-\alpha r}$$

Where A_0 is the amplitude of the source signal. Applying a logarithm to the equation, its become linear in α. The attenuation tomography is inverted using the previous results of seismic and using the W matrix as a gradient. We performed 5 iterations starting from a simple homogeneous media with an attenuation $\alpha = 1.0\,e^{-03}$ m-1, computing the Fresnel weights matrix for 5 different frequencies (30 Hz, 45 Hz, 60 Hz, 90 Hz, and 120 Hz). In order to link the attenuation map with surface cracking, we made an inventory of the fissuring state of the soil along the seismic profile. We kept all the cracks wider than 5 cm and created a Surface Cracking Index (SCI) corresponding to the total length of cracks per linear meter. The SCI is then normalized (Fig.5).

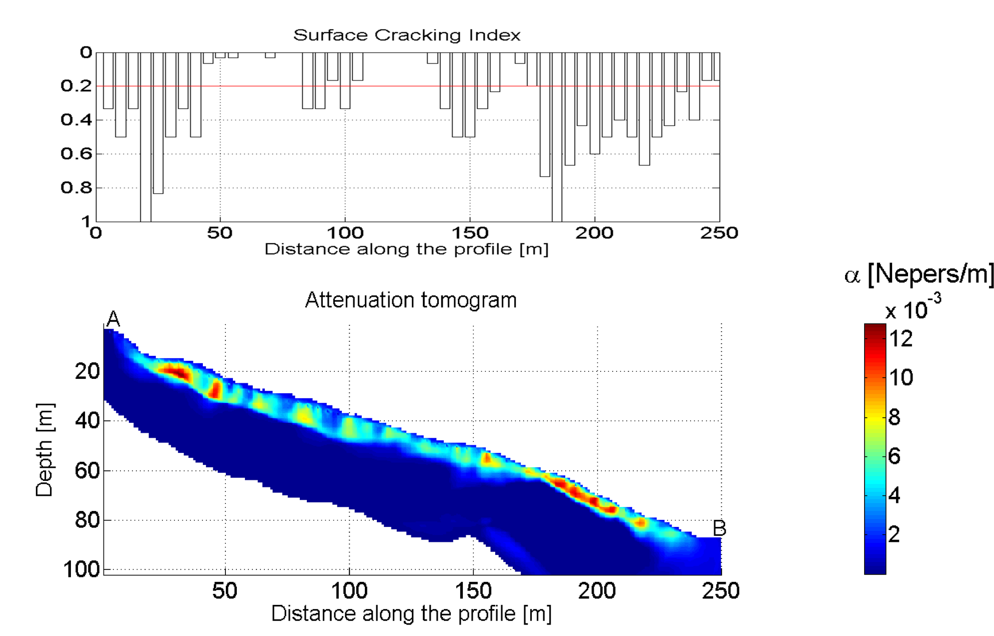

Figure 5 : Inversed attenuation tomography

Interpretation

Results from seismic P-wave inversion have been compared to dynamic penetration tests and to the 3D geological model of Travelletti and Malet (2011). The bedrock depths are comparable and the differences are below 5 meters. We also used an orthophotography and a DEM of the area from 1956 before the landslide triggering to compare the paleotopography recovered by seismic and the old topography. We notice on the orthophotography a first slight bump of the topography followed by a depression and a second important bump created by a crest crossing the seismic profile as well as on the P-wave tomography, respectively at 90, 125 and 140 m. Near 180 m, one can see a zone of low velocities (< 900 m/s) in depth that we interpret as the path of a gully present before its covering, also visible on the orthophotography of 1956 (Fig.6).

Bedrock and landslide layer are well recovered by seismic attenuation and results are well correlated to surface cracking. The values obtained for the bedrock are around 10-3 Np.m-1, which corresponds to consolidated soils attenuation at 40 Hz. The attenuation of the landslide layer varies from 4 to 12 10-3 Np.m-1, which are more characteristics of unconsolidated soils at that frequency. The lateral heterogeneity of the attenuation is important. Areas of high attenuation show attenuation 3 times higher than others. The correlation between the SCI and the attenuation is very good, so that we can affirm that attenuation variations are mainly caused by cracking. Results obtained from seismic data are gathered inside an interpreted model (Fig 6. b)) that contains the three units visible in the P-wave tomography and cracks area determined by high attenuation zones. This interpreted geological model could improve the result in numerical modeling for a better hazard assessment.

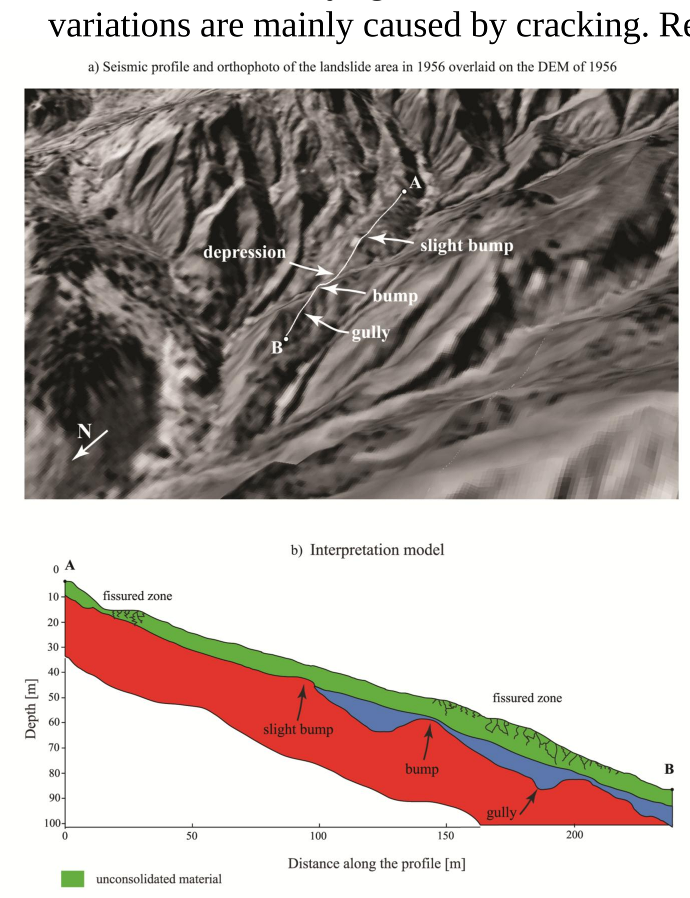

Figure 6 : a) geomorphological context of the earthflow and b) final cross-section interpretation

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